SEDIMENT YIELD FROM GLACIO-LACUSTRINE CALCAREOUS DEPOSITS DURING THE POSTGLACIAL PERIOD IN THE COMBE D'AIN (JURA, FRANCE)

M. CAMPY*, J. F. BUONCRISTIANI AND V. BICHET

Centre des Sciences de la Terre, UMR CNRS 5561, Université de Bourgogne, 6 Bd Gabriel, 21000 Dijon, France

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ABSTRACT

The middle valley of the river Ain (Jura) cuts through glacio-lacustrine deposits laid down in an ice-dammed lake during the most recent glacial advance. The total volume eroded is about $6{\cdot}21\times10^8$ m^3 for a surface area of $3{\cdot}7\times10^7$ m^2. Erosion occurred between 18 ka BP and 6 ka BP, i.e. over a duration of some 12 ka. Sediment yield from the area was of the order of 2500 t km^{-2} a^{-1}, which is comparable with modern-day sediment yield from NW African badlands. These high values are ascribed to the amenability of glacio-lacustrine deposits to mechanical weathering and to the rapid geomorphological changes that affected glacial and paraglacial sedimentary cover after the retreat of the ice. The valley slopes were destabilized by mass wasting (earthflow and mudflow), which was the predominant erosional process. The slopes are currently stabilized or very exceptionally active. © 1998 John Wiley & Sons, Ltd.

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KEY WORDS: quantification of erosion; glacio-lacustrine sediments; postglacial period.

INTRODUCTION

The magnitude of sediment yield and the speed of denudation of continental areas are fairly well documented for the present time. Works by Milliman and Meade (1983) and Meybeck (1984, 1987) evaluated yields on the basis of material reaching the oceans as measured at the estuaries of the world's major rivers.

Sediment yield is estimated at between 10 t km^{-2} a^{-1} (corresponding to a denudation rate of about $0{\cdot}01$ mm a^{-1}) for low-lying shield drainage systems (Canadian and Siberian shields) and more than 1000 t km^{-2} a^{-1} (denudation rate on the order of 1 mm a^{-1}) for the Himalayas and the East Indies. The wide range of values is brought about by variability in local conditions and in particular by three main parameters: rock type, relief and climate.

Studies have been conducted on smaller drainage basins where conditions are more uniform in an attempt to differentiate the effect of each parameter on the intensity of erosional processes (Hornung *et al*., 1990). The lithological characters (compaction, induration, fracturation, fissuration, etc.) of the substratum, for example, markedly affect the processes and intensity of mechanical weathering in particular (Meybeck, 1987; Einsele, 1992). Sediment yield from 153 small drainage basins in North Africa is put at 1000–2000 t km^{-2} a^{-1} for lightly compacted mudstones (badlands), 500–1000 t km^{-2} a^{-1} for marly limestones, 200–800 t km^{-2} a^{-1} for shales and pelites, 100–500 t km^{-2} a^{-1} for limestones and micaschists and 10–100 t km^{-2} a^{-1} for hard sandstones and granites (Probst, 1992). Drainage basin topography and prevailing climate account for the spread of values for each rock type. One of the highest recorded values in the world for sediment yield is for the Yellow River (Huang Ho) basin in China where the loess surface cover is particularly susceptible to erosion. Measurements over 40 years indicate sediment yields in the order of 1400 t km^{-2} a^{-1} from the entire basin (Ying Wang *et al*., 1986) and local yields of up to 20 000 t km^{-2} a^{-1} in smaller sub-basins (Chen *et al*., 1989).

All these data are based on measurements made in the last 30 years and therefore include erosion caused by human activities which is difficult to separate from natural erosion. Extrapolation from current data to earlier times is therefore unreliable (Berner and Berner, 1987; Meade, 1969, 1988; Einsele, 1992). Proposed

* Correspondence to: M. Campy

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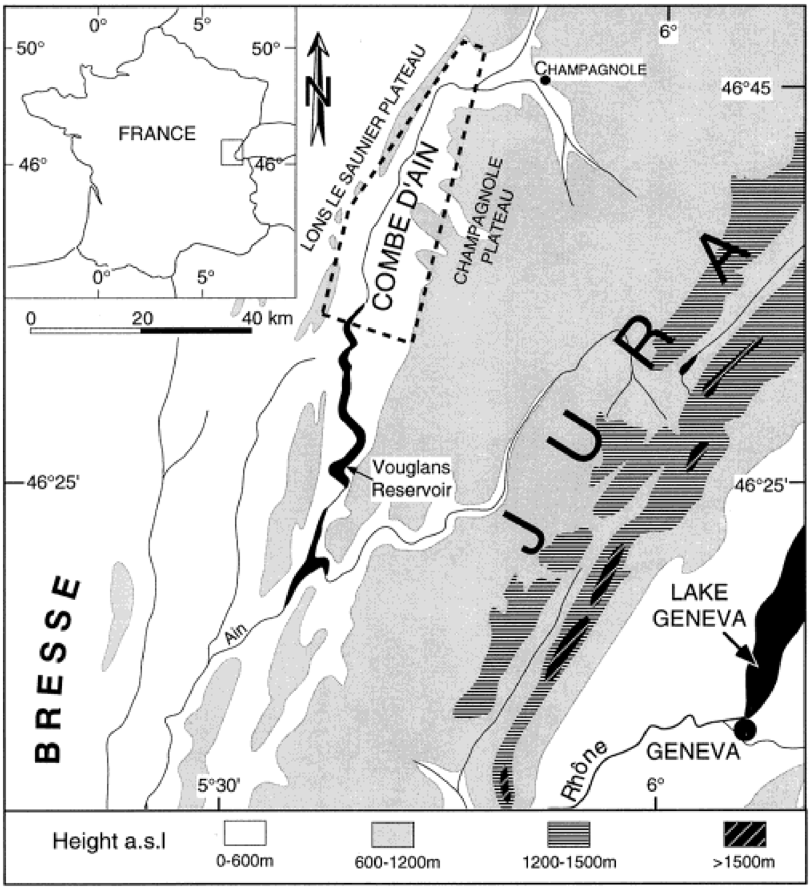


Figure 1. Location of the Combe d'Ain in the Jura range

extrapolations (Tardy *et al*., 1989; Probst, 1991) can only be of very broad relevance as they gloss over the diversity of individual cases and local situations.

Ideally, to make accurate estimates of sediment yield during the geological past one needs to be able to calculate the volume of rock eroded during a given interval of time (Einsele, 1992, p. 376). Suitable situations for this method are rare though. Mills (1976) used the method to calculate a mean denudation rate of 1100 mm ka^{-1} for the slope of 4000 m high andesitic Mt Rainier during the last 0·32 Ma. Similarly, Rice (1980) estimated denudation of Permian–Triassic mudstones, siltstones and sandstones in the Little Colorado valley of the order of 80 to 100 m Ma^{-1}, from the downcutting of the trunk river through radiometrically dated deposits of early Pliocene volcanic extrusions. In southern Germany, relics of Upper Jurassic limestones which formed the ancient landscape and have sunk into a Miocene volcanic pipe indicate that the land surface has been lowered by about 600 m during the last 18 Ma (Geyer and Gwinner, 1986). These data point to an average denudation rate of 33 m Ma^{-1}.

In this paper we investigate a special case where sediment yield related to mechanical erosion can be evaluated in glacio-lacustrine carbonate siltites during postglacial times.

GEOLOGICAL SETTING

The Combe d'Ain is a generally N–S oriented depression gouged out of the limestone plateaux of the western side of the Jura Mountains (Figure 1). It is bounded to the west by the plateau of Lons-le-Saunier and to the east

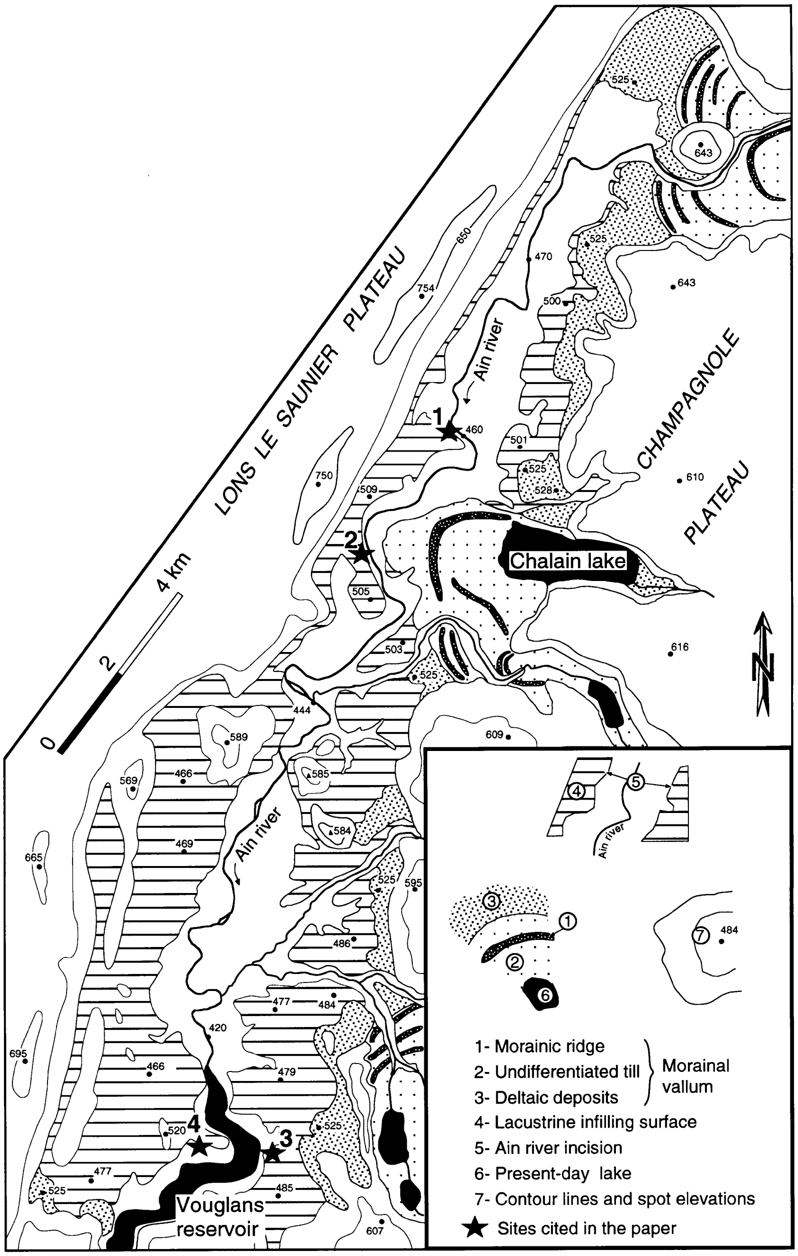


Figure 2. Simplified geological map of glacial and glacio-lacustrine formations of the Combe d'Ain

by the Champagnole plateau. The Combe is drained axially by the river Ain which was dammed in 1968 to form Vouglans reservoir.

The Combe d'Ain is partly filled by glacial and glacio-lacustrine formations (Figures 2 and 3) (Campy, 1982). On the eastern side three major terminal moraines and deltaic deposits (vallum) were left at the outlets of valleys cut into the Champagnole plateau. They mark the maximum advance of the Jura glacier during the last Ice Age (Campy, 1983). In contact with the moraines on the western edge of Champagnole plateau is a virtually unbroken terrace at 525 m asl composed of coarse delta deposits with classical foreset and topset beds of a Gilbert-type delta (Gilbert and Shaw, 1981). A terrace varying in elevation from 466 m to 509 m asl occupies the remainder of the Combe d'Ain. It is slightly higher on the eastern edge than on the western rim of the Combe. It

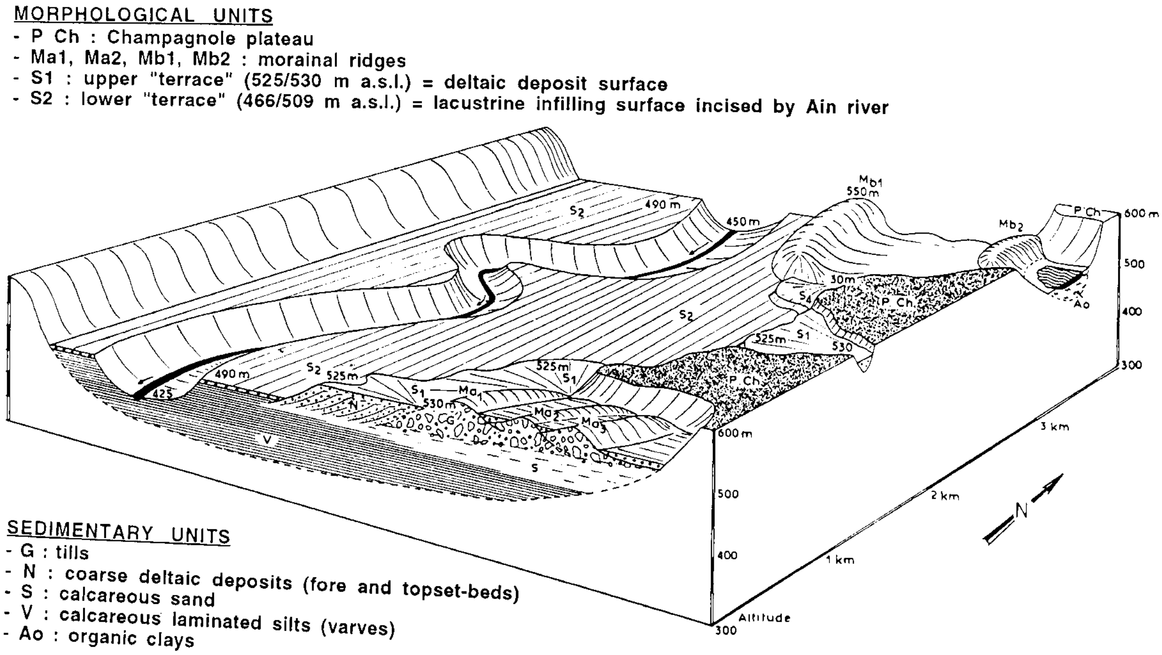


Figure 3. Block diagram of the morpho-sedimentary units of the Combe d'Ain

is made up of laminated silt–clay sediments (rhythmites or varves) characteristic of distal glacio-lacustrine sedimentation (bottomset bed) (Cohen, 1979).

This glacio-lacustrine formation has been downcut by the river Ain to a depth of about 30 m upstream and 60 m downstream relative to the top of the varve formation (Figures 3, 4, 5.1 and 5.2). The aim of this work is to calculate the volume of deposits eroded and the corresponding sediment yield.

CHRONOLOGICAL DATA

Origin and age of the lacustrine infilling

During the last glacial maximum (Würm) the Jura Mountains were covered by an ice cap that gave rise to a number of peripheral ice tongues (Campy and Arn, 1991). These tongues dammed certain depressions causing proglacial lakes to form (Figure 6). Melt waters deposited their detrital load in these lakes before draining off northwards as the Doubs and southwards as the Ain. The map of glacial and proglacial deposits (Figure 2) can be used to reconstitute the palaeogeography of the glacial margin at the time of the maximum advance in the Combe d'Ain (Figure 7). Combe d'Ain lake was dammed to the south by the Orgelet ice-tongue. The lake level remained stable for some time at about 525/530 m asl (altitude of the proglacial delta topset bed). Four ice-tongues fed the lake.

- The Champagnole ice-tongue to the north terminated short of the lake allowing the formation of a broad proglacial delta.
- The Doucier ice-tongue in the centre jutted into the lake so no classical end delta could form but only fans on both sides. Waterlain tills were deposited in front of the ice-tongue (Dreimanis, 1979; Campy, 1983).
- The Clairvaux ice-tongue terminated back from the lake (cf. Champagnole) allowing a wide proglacial delta to form.
- The Orgelet ice-tongue to the south extended into the lake (cf. Doucier ice-tongue) depositing waterlain tills.

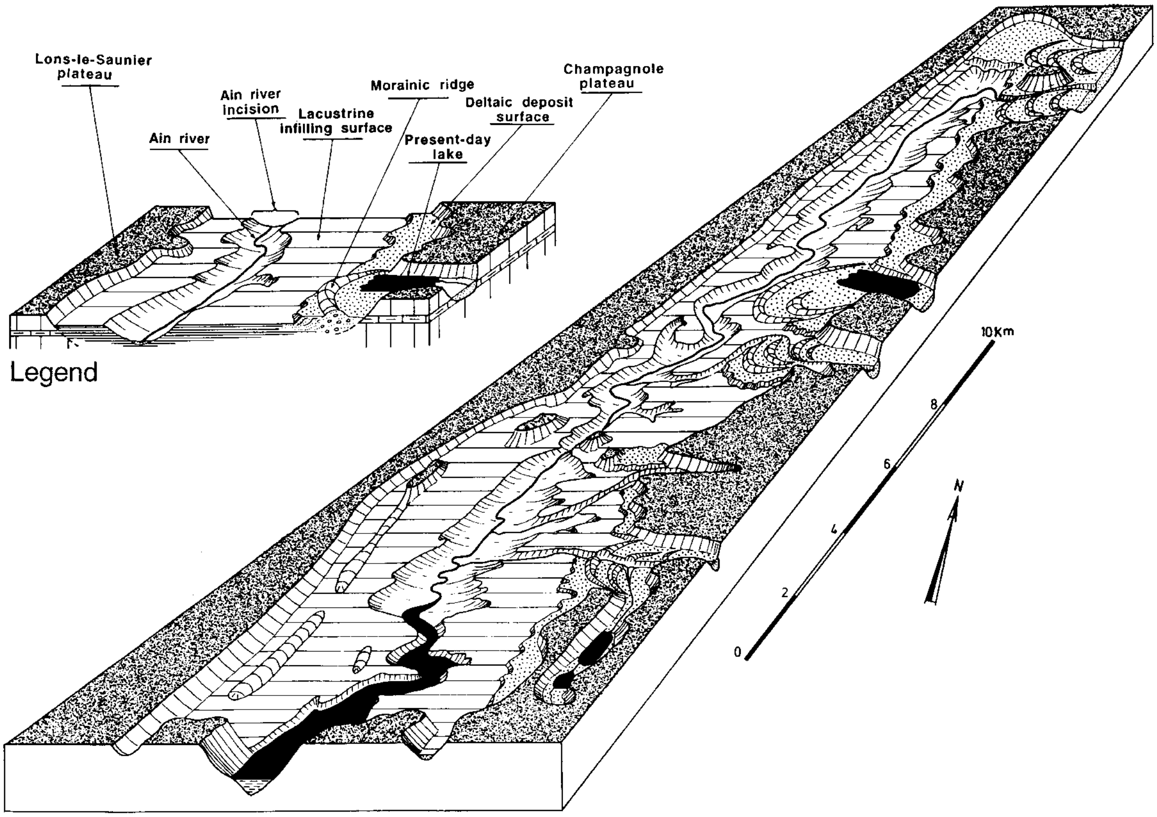


Figure 4. Block diagram of the Combe d'Ain showing how the river has cut through the varve deposits

A count of the varves deposited in the Combe d'Ain shows that the glacio-lacustrine system was operative for about 1000 years (Campy, 1982).

The glacial maximum is difficult to date as no fossils or organic remains have been found in the corresponding sediments. However, the maximum has been dated indirectly through the impact on proglacial deposits close to the front and on periglacial deposits on the eastern edge of the Jura (Campy and Richard, 1988; Campy and Chaline, 1993). The onset of glacial advance is dated to 21/22 ka BP, with the maximum at between 19 and 18 ka BP and retreat from 18 ka BP.

Mode and age of the onset of erosion

The good state of conservation of the proglacial deltas and the absence of secondary deltas are evidence that the lake emptied in a single relatively short event. As the glacier began its retreat, the Orgelet ice-tongue undammed the lake and its waters drained southwards into the ice-free area. A N–S drainage axis rapidly became established at the surface of the emerged lake-bed deposits. Locally these deposits are overlain by up to 1 m of gravel marking the beginning of the postglacial course of the Ain. The gravel probably derives from gullying of the coarse deltas that stood 20 to 30 m above the level of the new river. We think the river began downcutting through the varve deposits as soon as the lake had emptied, which was probably a rapid process (several decades?) around 18 ka BP.

Age and duration of the erosional phase

Erosional processes are not very active at present in the Combe d'Ain. Estimates of suspended load made by reservoir supervisors at Vouglans (oral report) where the Ain flows into the reservoir are of the order of

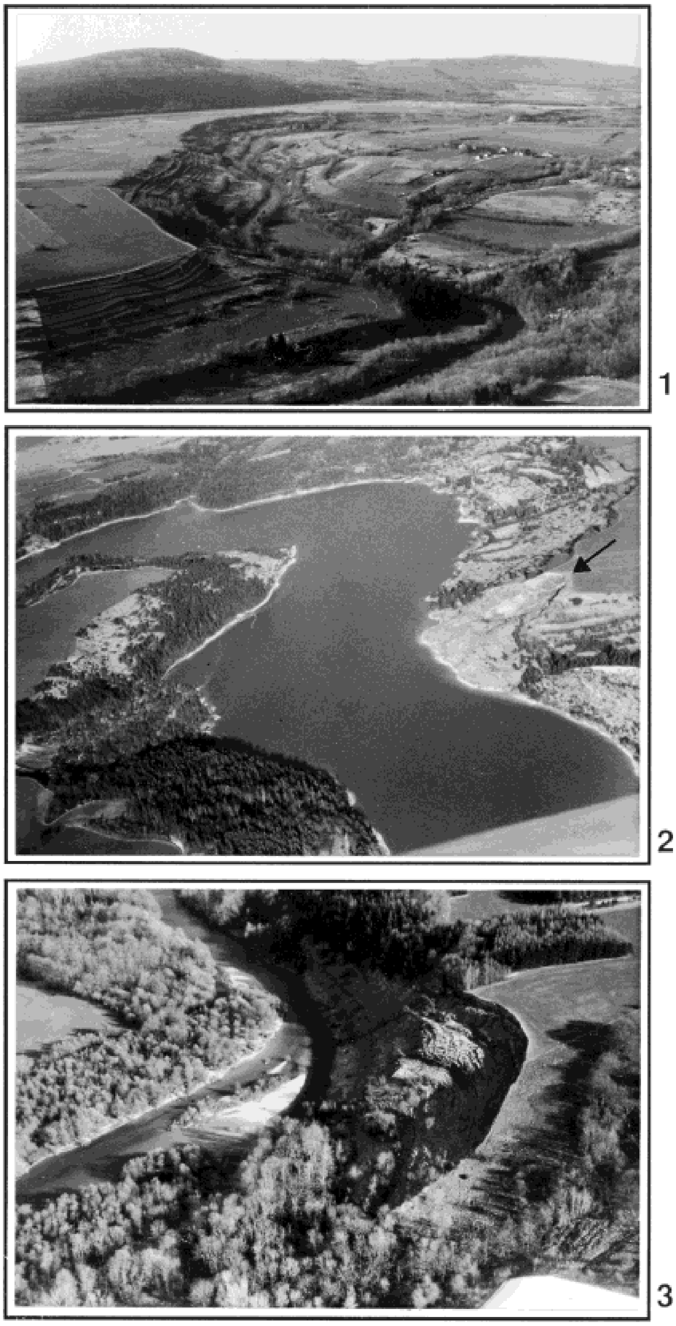


Figure 5. Ain river incision and present-day landslide in the Combe d'Ain. (1) View of the Ain valley cut into the varve deposits (Figure 2, site 2). The lacustrine infilling surface can be seen on the left and the edge of the Lons-le-Saunier plateau in the distance. The man-made terraces on the hillslopes follow the scars of ancient landslides. (2) View of the Ain valley incision currently occupied by Vouglans reservoir (Figure 2 between sites 3 and 4). The lacustrine infilling surface can be seen on either side of the incision. The valley sides are currently stable. A small mudflow associated with a small spring that locally wets the varve deposits is arrowed. (3) Ébalèves landslide site (Figure 2, site 1). This is the only earthflow currently active in the Combe d'Ain. On the right is the lacustrine infilling surface. The river Ain flowing 40 m below it erodes the base of the varve deposits during floods causing the landslip. The head of the landslide is marked by a 3–4 m high shear scar and the hillslope is punctuated by flats that slope slightly upstream. These stair-like counter-slopes are the top of small blocks that collapsed towards the slope

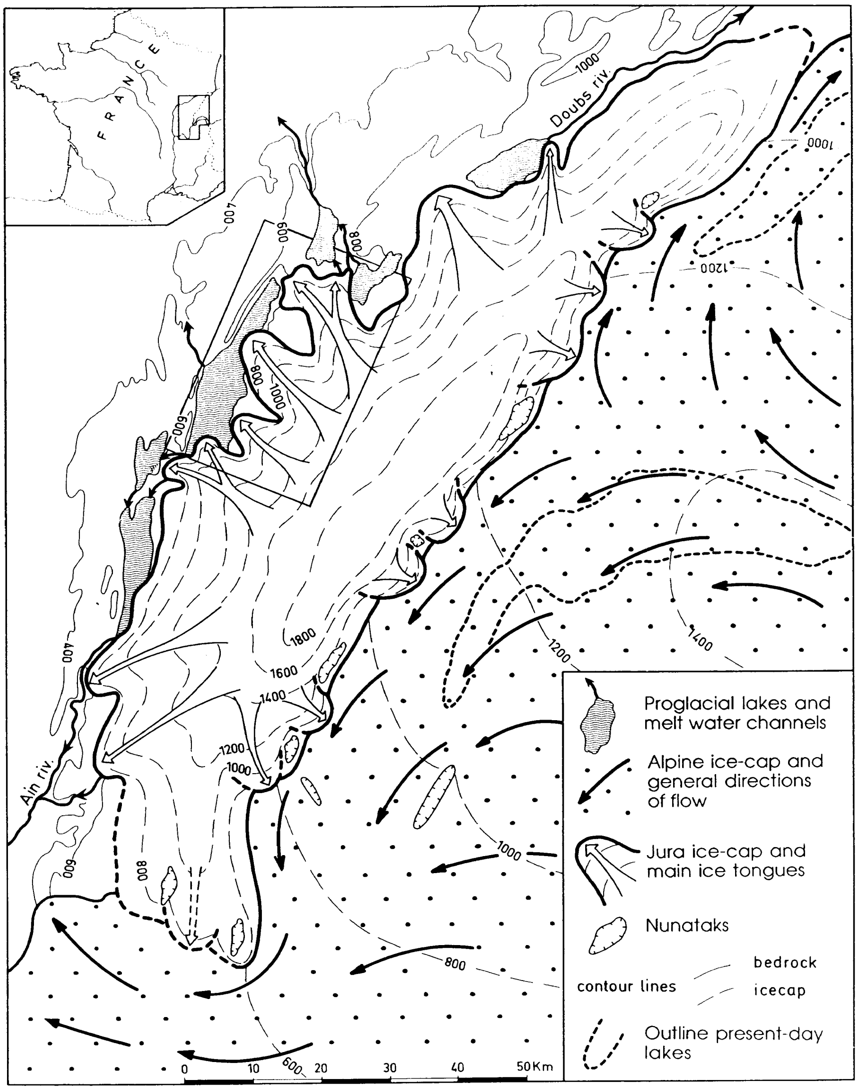


Figure 6. The Jura ice cap at the Würm ice age maximum (18 ka BP). Location of Figure 7 is outlined (modified after Campy and Arn, 1990)

10–15 t a^{-1}. The dam was built in 1968 and surveys of the reservoir bottom and upstream area reveal no visible detrital cone. The sides of the valley that has been cut into the varve deposits seem to be stabilized at present throughout the Combe d'Ain area. Scars of earlier landslides are visible as they have been transformed into terraces by human activity (Figure 5.1).

Mechanical erosion is currently active at a few localized points only.

- In the Vouglans reservoir area a mudflow can be seen where there is local input of throughflow water (Figure 5.2).
- Upstream, the Ain flows generally along the valley centreline well clear of the hillslopes. Only at one spot (Ébalèves site) does the river form a meander eroding the hillslope (Figure 2, site 1 and Figure 5.3). Here a complex landslide currently functions as an earthflow associated with mudflows. We report little activity in 30 years of observation and the mass of material mobilized is estimated at some 5–10 m^3 a^{-1}.

It seems therefore that apart from these rare areas, the valley sides are currently stable and mechanical erosion is very slight. The question is when did downcutting begin and subsequently abate? There is one clue to the riddle.

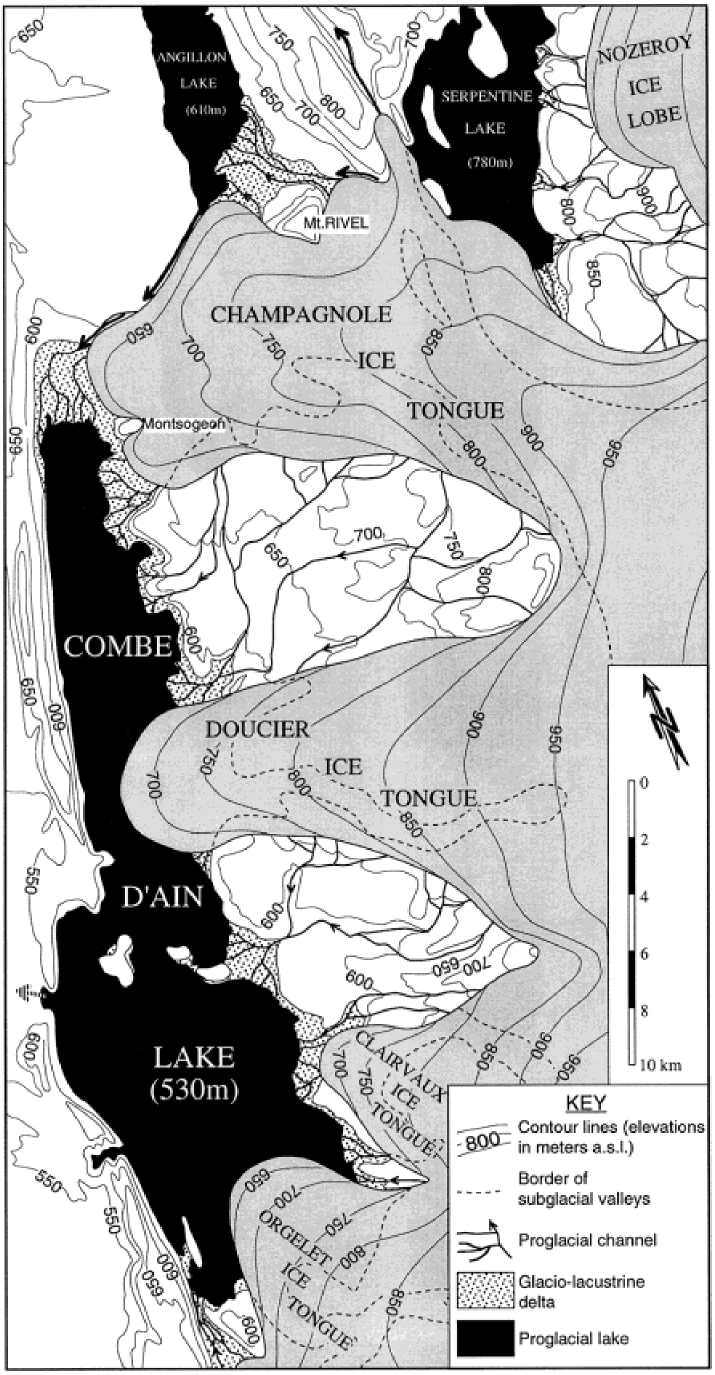


Figure 7. Palaeogeography of the glacial margin during the Würm ice age maximum in the proglacial lake of the Combe d'Ain (modified after Campy, 1982)

On the western edge of the Vouglans reservoir, at the top of a stabilized ancient landslip (Figure 8, Figure 2, site 4 and Figure 9.1) a peat lens was found enclosed in lacustrine deposits. Palynological analysis (Dr J. Heim, University of Louvain-la-Neuve, Belgium) of 350 pollen grains revealed 84 per cent of tree pollen (*Tilla* 65 per cent, *Pinus* 15 per cent, *Corylus* 4 per cent) and 16 per cent of thermophilous herbaceous plant pollen (grasses)

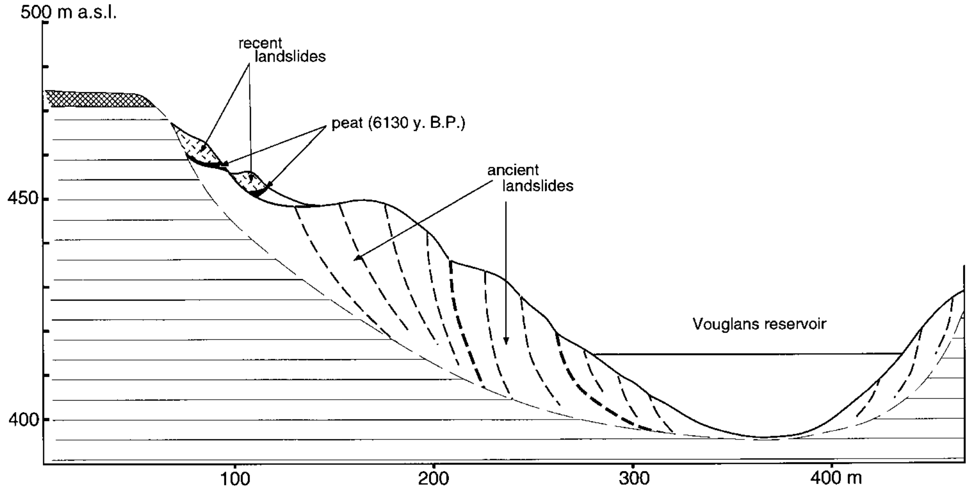


Figure 8. Cross-section of the ancient landslide site at Épiney (Figure 2, site 4). The peat lens dated to 6130 BP (Figure 9.1) overlies the ancient landslide and was preserved beneath a small recent landslide marking the end of major activity on the hillslope

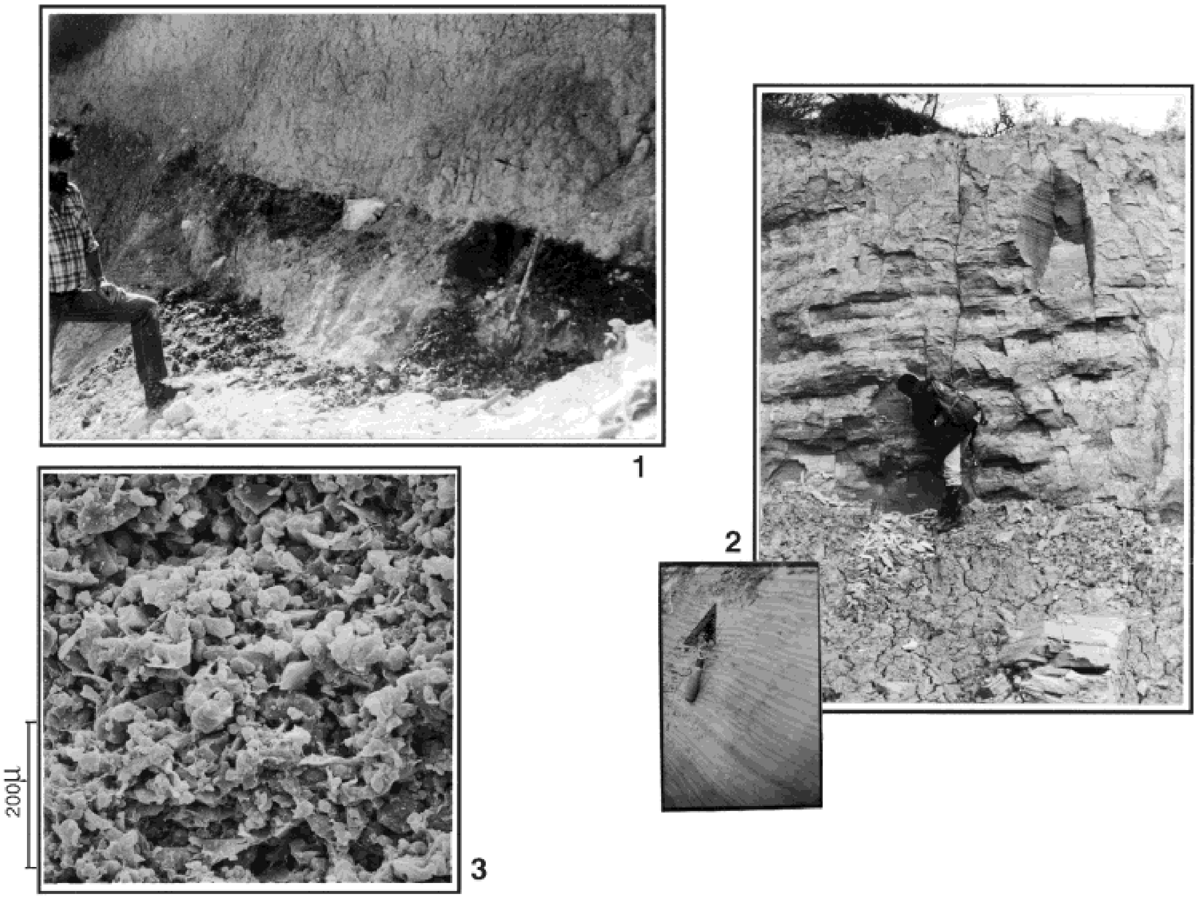


Figure 9. Lacustrine infilling deposits in the Combe d'Ain. (1) Peat lens at the Épiney site (Figure 2, site 3). See main text for commentary and Figure 8 for stratigraphic location. (2) Varve formations in the Ébalèves landslide (Figure 2, site 1). (3) Scanning electro-micrograph of glacio-lacustrine siltites (see text for commentary)

and fern (*Polypodium*) and horsetail (*Equisetum*) spores. The palynological assemblage is typical of lime forests with scattered pines and hazel undergrowth. The herbaceous stratum is dominated by ferns. The flora is characteristic of the Atlantic period in the Jura (Richard, 1983). Carbon-14 dating (LV – 1901) of the wood in the peat lens indicated an age of 6130±90 years BP, which sets deposition of the lens within the recent Atlantic climatic period and confirms the palynological data. The lens formed in a small bog lodged in the scar of an earthflow and was subsequently buried by a final landslide (Tallis and Rowntree, 1980). The position of the lens shows that the hillslope has been virtually stable since at least 6000 years BP. This is consistent with previous findings (Campy *et al*., 1994) about the Holocene infilling of Jura lakes, where the input of clastic deposits was curtailed during the Atlantic period when hillslopes were thickly wooded. It can be considered then that the Ain valley was cut between 18 ka and 6 ka BP. The process therefore lasted 12 ka.

INFILLING MATERIALS

Petrographic nature

The present-day Ain valley is cut through two types of glacio-lacustrine materials: clay–silt rhythmites and waterlain tills.

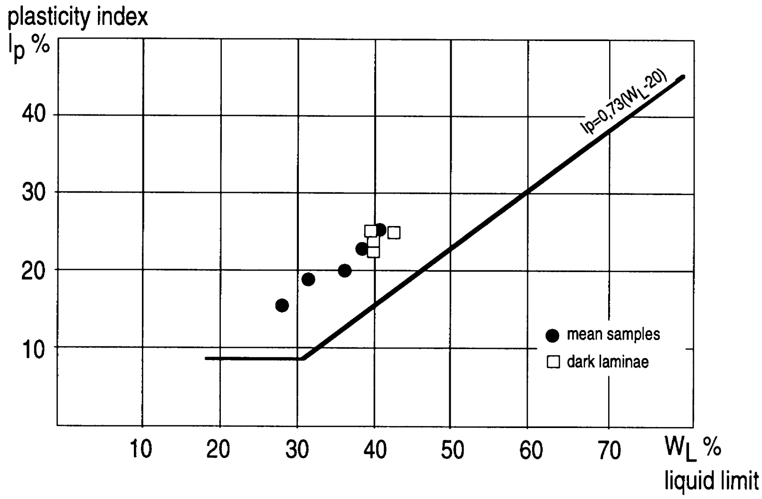
The main characteristic of rhythmites is their alternating light and dark pattern of horizontal centimetric banding (Figure 9.2). They include isolated blocks assumed to be dropstones. The rhythmites are stacked more than 40 m thick in the centre of the Combe d'Ain and are coarser (silt–sand) on the eastern edge of the Combe than on its western edge (clay–silt). Calcium carbonate is the principal constituent (85–98 per cent of 90 measurements) (Campy, 1982). The small proportion of clay is 60 per cent illite and 35 per cent smectite, with trace amounts of kaolinite (Biot *et al*., 1981). Analysis reveals no organic matter in either the pale or the dark laminae. The difference in colour is due to the difference in grain size distribution. The mean grain size is close to 30 µm in the pale bands and 5–10 µm in the dark ones (Campy, 1982). Scanning electron microscopy (SEM) (Figure 9.3) shows that the particles form a jumbled mass with many pore spaces. Calcite crystals can be identified although the particles do not usually exhibit distinctive crystal habits. They seem to be derived mainly from limestone rock flour eroded from valley slopes by the glacier. This banded formation is composed primarily therefore of clastic material as would be expected for glacial deposits.

Waterlain tills also occur in the Combe d'Ain infilling but in smaller amounts. They are found only in the areas where the Doucier and Orgelet glacial tongues jutted into the proglacial lake (Figure 7). The tills differ little in composition from the rhythmites discussed previously. Two characteristics stand out: more abundant cobbles and boulders (10–20 per cent) scattered through the silt matrix, and creep structures showing that the deposits were laid down subaqueously. The abundant matrix of these waterlain tills means they have the same geotechnical characteristics as the foregoing rhythmites.

Geotechnical data

The rhythmite formation has a dry specific gravity of the order 1·8 t m^{-3} for a mean intrinsic porosity of 30 per cent. Moisture content (W) of outcrop samples in the natural state varies from 20 to 26 per cent depending on lamina grain size. These values are very close to saturation and confer high moisture sensitivity on the formation. Shallow saturation (0·5–1 m) occurs rapidly and wet weather commonly causes waterlogging of overlying meadows.

The mechanical uniformity of the rhythmites and their variability in behaviour versus moisture content were tested in the laboratory using the Atterberg Limits method (LCPC, 1970) and in the field with the Soiltest penetrometer test. The Atterberg Limits were measured on nine samples from different outcrops of either a set of laminae or a selection of dark bands. The results plotted on a plasticity diagram (Figure 10) indicate variation in the liquid limits (W_L – consistency limit between plastic and liquid) of between 29 per cent and 41 per cent. The plasticity index ($I_p = W_L - W_p$, where W_p is the consistency limit between solid and plastic states) varies correlatively between 15 per cent and 24 per cent. These values indicate the mean plasticity of the rhythmites and reflect the fairly uniform behaviour for the set of samples as a whole. The dark bands produce the highest W_L values (40–42 per cent). *In situ* penetrometer tests were carried out on parallel, perpendicular and oblique



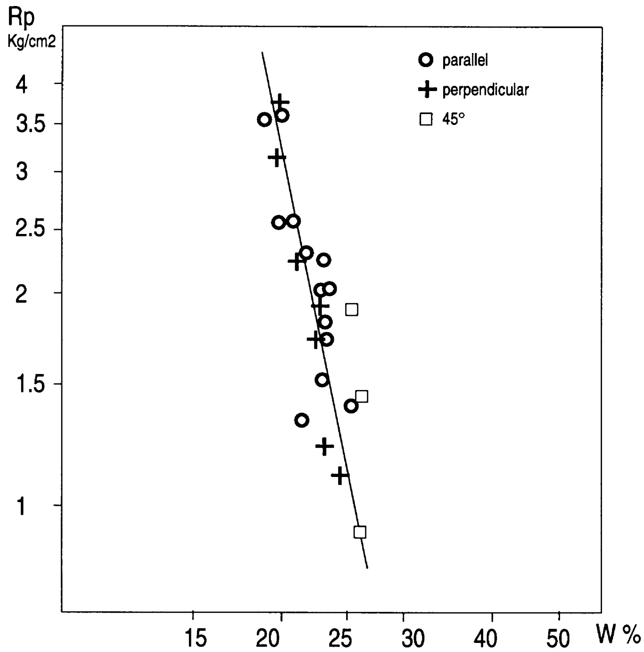


Figure 10. Geotechnical tests on varve deposits. Top: plasticity diagram (Atterberg limits). Bottom: moisture content versus penetration resistance

surfaces (45°) relative to the lamina plane. Penetrometer resistance values (R_p) obtained for 23 tests were correlated with water content (W) on a semi-logarithmic chart (Figure 10). It can be seen that:

- resistance to penetration is not markedly influenced by the direction of the force relative to the lamina plane;
- there is a close correlation with moisture content;
- the mechanical strength of the formation is uniform for all the outcrops tested.

These summary geotechnical tests complete the sedimentological data. They reveal uniform behaviour of the rhythmite formation, despite its laminated structure. They emphasize the porous nature of materials already observed by SEM and their plasticity.

The marked moisture sensitivity of the facies which are easily saturated or supersaturated and their low intrinsic cohesiveness explain the instability of the hillslopes which allowed the river to erode through this formation.

Process and nature of erosion

Earth movements that are still active at some places in the Combe d'Ain are indicative of the processes that led to the incision of the valley: earthflows and mudflows.

The active slump at Ébalèves (Figure 5.3) is typical of saturated or supersaturated loosely cohesive sediments. The slump head is marked by a 3 to 4 m shear scar and the hillslope is punctuated by flats that are each the top of small blocks that collapsed towards the slope. The ancient landslide that is currently stabilized (Figure 8, Figure 2, site 4) on the west bank of Vouglans reservoir (Épiney site) is a further example of the processes that affected the hillslopes as the Ain valley was cut.

A mudflow on the hillslope above Vouglans reservoir (Figure 5.2 and Figure 2, site 4) is activated in some years during winter or early spring by heavy rainfall or melting snow. It is associated with a small spring that locally moistens the varve deposits.

Cutting of the Ain valley since the lake dried up must essentially have involved these two processes of mechanical erosion (earthflow and mudflow). Although negligible at present in terms of the volume of erosion, the processes were far more active and commonplace between 18 and 6 ka BP.

ERODED VOLUME AND SEDIMENT YIELD

The total volume of sediment eroded by the Ain was located between the present-day valley surface (Ps) and the initial surface (Is) of the lake infilling. The eroded area corresponds to the intersection of Is and Ps (Figure 11.1).

Method

The volume and area of sediment eroded by the river Ain were calculated by mathematical modelling of surface areas Ps and Is (Surfer software). The bounds of the incised valley were defined from the IGN 1:20 000 topographical map and supplemented by field observations. To facilitate calculation of the volume, the incision was divided into five morphologically similar zones (Figure 11.2). Each zone was processed separately for the different stages of the calculation and the final result is the sum of the five partial results.

The surface areas were mathematically modelled by transforming the points measured on the topographical map into points calculated by interpolation at each node in a grid defined for the model. The Kriging method of interpolation was selected (linear variogram) which uses the theory of regional scale variables (Guillaume, 1977). To avoid error due to interpolation, enough points had to be measured to provide a uniform spread over the entire study zone. To comply with these constraints the points were chosen by using a 200×200 m standard grid. The points on the valley sides correspond to the points of intersection of the grid and the contour lines; in flat areas they lie at the nodes of the standard grid. Because the contours are evenly spaced (10 m) we estimate an error of 5 m for the values measured.

After defining the bounds of surface areas S_i and S_a, each zone was processed as follows:

- selection of measurement points (Figure 11.3A) using a 200×200 m grid and digitalization of the X, Y, Z coordinates;
- interpolation by the Kriging method (Figure 11.3B) to calculate X′, Y′ and Z′ from the measured X, Y, Z values;
- modelling (Figure 11.3C) from the computed coordinates (X′, Y′, Z′) in two dimensions (maps) and three dimensions (block diagrams).

Results

The volume between the modelled surface areas (Ps and Is) corresponds to the eroded volume. It is calculated by the integral method (Simpson) with Surfer software. The calculated volume of eroded sediment is

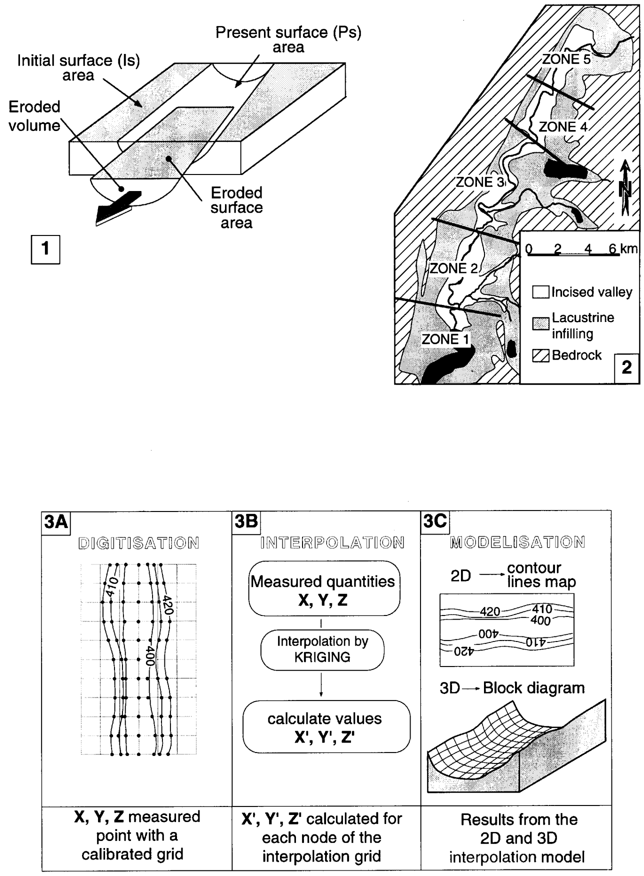


Figure 11. Calculation of the volume of eroded material. (1) The different surfaces modelled used for the volume calculation. (2) Location of the five zones whose volumes were calculated separately. (3) Processing carried out for each of the five zones

$6{\cdot}21\times10^{8}$ m^3. An average uncertainty of 15 per cent on the calculation of the volume can be estimated from the $\pm2{\cdot}5$ m error on the measurements.

The eroded area calculated by the same method is $3{\cdot}7\times10^{7}$ m^2.

Sediment yield in cubic metres per year (Q_e) is calculated using the formula:

$$Q_e = V/t$$

where V is total volume of sediment eroded ($6{\cdot}21\times10^{8}$ m^3) and t is duration of erosion (12 000 years). Sediment yield is 51750 ± 7700 m^3 a^{-1}.

Sediment yield in tonnes per year (Q_m) is calculated using the formula:

$$Q_m = Q_e \times d$$

where d is mean sediment density ($1{\cdot}8$ t m^{-3}). Sediment yield is 93150 ± 13970 t a^{-1}.

Sediment yield in tonnes per year per kilometre square (Q_s) is calculated from the formula:

$$Q_s = Q_m/S$$

where S is eroded area ($3{\cdot}7\times10^{7}$ m^3). Sediment production is 2520 ± 380 t km^{-2} a^{-1}.

CONCLUSION AND DISCUSSION

The mean sediment yield of about 2500 t km^{-2} a^{-1} during 12 000 years is high compared with values obtained over very short periods for the present day and referred to in the introduction to this paper. It is close to present-

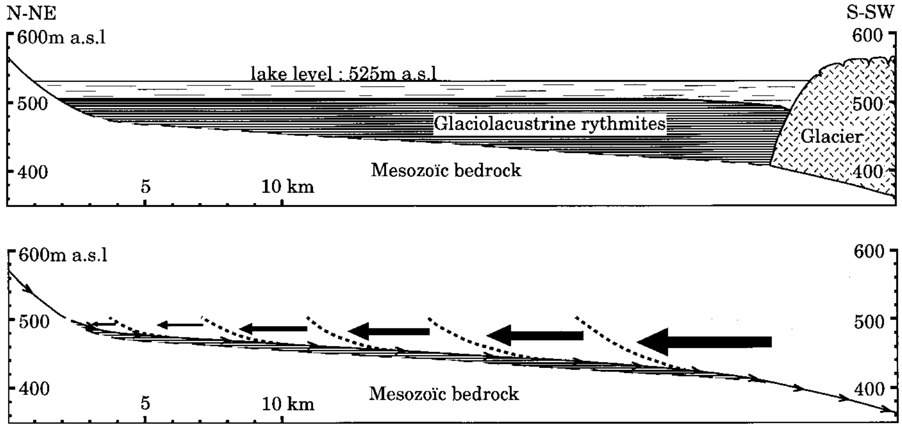


Figure 12. Longitudinal cross-section of the Combe d'Ain. Top: at the end of the lacustrine infilling (18 ka BP). Bottom: gradual erosion of varve deposits from 18 to 6 ka BP

day values recorded for small badlands-type drainage basins on mudstones in NW Africa where the terrain is very uneven and the rainfall regime highly irregular. It is comparable with values for the Huang-Ho basin which is currently much affected by farming techniques (Einsele, 1992, p. 356). Among the three parameters that usually affect mechanical erosion (relief, climate and rock type) it is lithology that seems to be the main factor behind the high values in the case studied. The relief is low, the climate is continental with fairly low temperature and rainfall variations during the postglacial period (Guiot, 1987; Zagwijn, 1994). The lithological characters of the glacio-lacustrine carbonate silts of the Combe d'Ain and in particular their weak intrinsic cohesiveness and their amenability to moisture saturation explain why they were so prone to mechanical erosion.

The findings of this work allow us to present the state of the Combe d'Ain before (*c.* 18 ka BP) and after the erosion stage but do not allow us to present the pattern of erosion and the changing morphology of the Combe d'Ain during this period. We can advance a plausible answer to this puzzle. Figure 12 shows a schematic cross-section of the Combe d'Ain and its glacio-lacustrine infilling. As stated above, the Combe d'Ain proglacial lake was dammed by an ice-tongue and the infilling abutted against the tongue. The void formed when the ice-dam melted around 18 ka BP caused substantial gravitational imbalance in the downstream zone of the infilling and downcutting certainly began in this zone as a very steep ravine. Downcutting of the ravine extended gradually upstream until the present-day position. It may be thought that sediment yield was not constant from 18 to 6 ka BP for two reasons. Firstly the hillslopes were much steeper at the beginning than at the end of the period and the earthflow processes therefore more active. Secondly palynological analysis (Richard, 1983) shows that plant cover was sparse during late glacial times (18–10 ka BP), while forest species colonized the region progressively from 10 ka BP onwards establishing total cover during the early Atlantic period around 6 ka BP. The development of plant cover would certainly have slowed mechanical erosion. Figure 13 shows the likely pattern of change. The low rate of erosion at the present day is only a continuation in the slow-down of erosion that began at the beginning of the Holocene.

The example of the Combe d'Ain also invites more general considerations. Many workers (Pinet and Souriau, 1988; Tardy *et al.*, 1989; Probst, 1991; Einsele, 1992 pp. 378–384) suggest extrapolating from present-day average values for erosion to geological times to estimate the rate of denudation of the continents. This entails a risk of serious error as some authors have emphasized (Berner and Berner, 1987; Meade, 1969, 1988). Current data are based on measurements made in the last 30 years and therefore include erosion caused by human activity which it is difficult to differentiate from natural erosion. The Combe d'Ain example indicates that erosion rates have changed greatly since 18 ka BP. In addition, estimates of erosion during the geological past can only be of very general significance, masking the diversity of cases and local situations. The Combe d'Ain example shows that the high rates of erosion reported are only valid during a limited period and over a

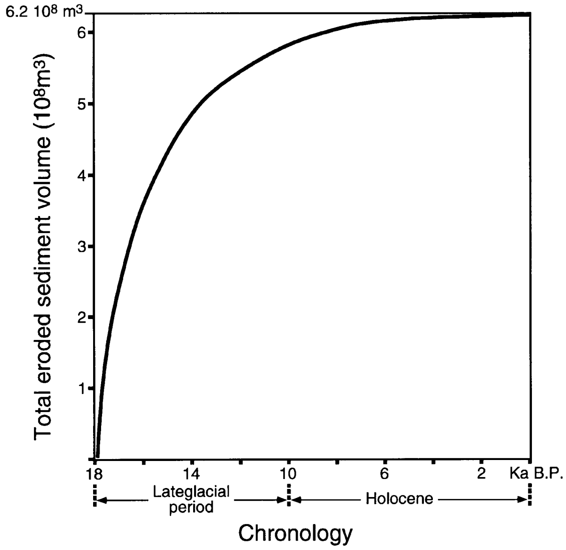


Figure 13. Hypothetical curve of the increase in the volume of sediment eroded over time

specific area of about 40 km^2 where carbonate silts crop out. Around this zone, on the limestone plateaux of the Jura, erosion is some 50 to 100 times weaker. These local variations in erosion rates account for present-day morphology.

The example of the Combe d'Ain also illustrates the scale on which mechanical erosion affected regions on the periphery of the last major glacial advance once the ice had retreated. Susceptibility to erosion results from the occurrence of abundant clastic cover left behind by glaciers. Such glacial and periglacial superficial formations (Church and Ryder, 1972) are usually uncompacted and topographically unstable. Glacial retreat causes intense modification of the processes of sedimentation and erosion. During the glacial phases *sensu stricto*, the sedimentation/erosion balance is largely positive. Conversely, glacial retreat opens the way to increased erosion and geomorphological adjustment at the expense of deposits that accumulated during the preceding phase. We have already reported a detrital 'rush' in the Jura in late glacial times (Campy *et al*., 1994) in postglacial lake infillings. The marked mobilization of clastic particles gradually abated during the Holocene as reliefs became more regular and vegetation took hold. This phenomenon probably occurred in all the peripheral areas of the most recent glacial advance.

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